Basin scale estimates of evapotranspiration using GRACE and other observations

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[1] Evapotranspiration is integral to studies of the Earth system, yet it is difficult to measure on regional scales. One estimation technique is a terrestrial water budget, i.e., total precipitation minus the sum of evapotranspiration and net runoff equals the change in water storage. Gravity Recovery and Climate Experiment (GRACE) satellite gravity observations are now enabling closure of this equation by providing the terrestrial water storage change. Equations are presented here for estimating evapotranspiration using observation based information, taking into account the unique nature of GRACE observations. GRACE water storage changes are first substantiated by comparing with results from a land surface model and a combined atmospheric-terrestrial water budget approach. Evapotranspiration is then estimated for 14 time periods over the Mississippi River basin and compared with output from three modeling systems. The GRACE estimates generally lay in the middle of the models and may provide skill in evaluating modeled evapotranspiration. *INDEX TERMS:* 1640 Global Change: Remote sensing; 1818 Hydrology: Evapotranspiration; 1836 Hydrology: Hydrologic budget (1655); 3337 Meteorology and Atmospheric Dynamics: Numerical modeling and data assimilation. **Citation:** Rodell, M., J. S. Famiglietti, J. Chen, S. I. Seneviratne, P. Viterbo, S. Holl, and C. R. Wilson (2004), Basin scale estimates of evapotranspiration using GRACE and other observations, *Geophys. Res. Lett.*, *31*, L20504, doi:10.1029/2004GL020873.

1. Introduction

[2] Evapotranspiration links Earth's water, energy, and carbon cycles. Roughly 50% of the solar radiation incident at the land surface is returned to the atmosphere as latent heat [*Kiehl and Trenberth*, 1997]. Evapotranspiration from the land surface replenishes atmospheric moisture and helps to sustain storms through the process of precipitation recycling [*Brubaker et al.*, 1993; *Eltahir and Bras*, 1996; *Bosilovich and Schubert*, 2002]. It also regulates the spatial-temporal distribution of soil moisture, which is itself a critical lower boundary forcing on climate [*Dirmeyer et al.*, 1999; *Koster et al.*, 2000]. Trends in evapotranspiration rates may be an indicator of climate change, in particular the acceleration of the hydrological cycle and changes in the way heat is redistributed from the tropics to midlatitude and polar regions [*Brutsaert and Parlange*, 1998; *Ohmura and Wild*, 2002; *Roderick and Farquhar*, 2002]. Therefore, improved characterization and quantification of evapotranspiration is essential for improving understanding of Earth system processes.

[3] Nevertheless, evapotranspiration is difficult to estimate at regional (climatic) scales. Micrometeorological measurement networks are generally too sparse for routine monitoring. Remote-sensing approaches typically rely on observations of surface temperature and vegetation indices as input to turbulent transfer or energy balance formulations. Several authors have demonstrated the strengths and weaknesses of these approaches [e.g., *Norman et al.*, 2001; *Kustas et al.*, 2001; *Jiang and Islam*, 2001]. Perhaps the most important limitation is the necessity of region-specific calibration using ancillary data such as air temperature, wind speed, surface resistance parameters, and/or independent estimates of latent and other heat fluxes.

[4] One approach to estimating regional evapotranspiration (ET) is solution of the drainage basin water balance for ET:

$$ET = P - Q - \Delta S, \tag{1}$$

where P is total precipitation, Q is net stream flow, and ΔS is the change in terrestrial water storage for a specific time period. *Yeh et al.* [1998] applied this method with reasonable success over Illinois, using in situ observations to estimate ΔS. However, while P and Q are often observed with sufficient accuracy to avoid large errors in the residual, independent estimates of ΔS have been lacking for most of the globe.

[5] The Gravity Recovery and Climate Experiment (GRACE) satellites, launched 17 March 2002, are now measuring Earth's gravity field with enough precision to infer terrestrial water mass variations over sufficiently large regions [*Wahr et al.*, 2004; *Tapley et al.*, 2004b]. Groundwater, soil moisture, snow, and surface water all may contribute significantly to the observed ΔS, yet GRACE alone can not separate them, making detailed hydrological interpretation a challenge [*Rodell and Famiglietti*, 2001, 2002]. Whereas ΔS estimates based on modeled or in situ data are apt to overlook one or more components, ΔS derived from GRACE is a perfect fit for water budget studies because it is a horizontally and vertically integrated quantity [*Rodell and Famiglietti*, 1999].

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[6] The objective of this paper is to demonstrate the estimation of *ET* using a terrestrial water budget approach with observation based ΔS, *P*, and *Q* estimates. The Mississippi River basin was chosen as the study region due to the availability of data, however, the method could theoretically be applied to any large drainage basin.

2. Data

[7] Through March 2004, GRACE had delivered 15 near-monthly global gravity field solutions, as sets of Stokes coefficients to a spherical harmonic expansion up to degree and order 120 [e.g., *Tapley et al.*, 2004a]. The effects of atmospheric surface pressure and ocean bottom pressure changes are removed using output from the European Centre for Medium-Range Weather Forecasts (ECMWF) operational forecast model and a barotropic ocean model driven by ECMWF pressure and winds [*Tapley et al.*, 2004a]. The variability in the resulting monthly gravity solutions is introduced mainly by redistribution of terrestrial water mass. The observed gravity signal degrades at higher degrees and orders (shorter length scales), hence there is a tradeoff between signal accuracy and precision in the delineation of a study region. As the selected minimum averaging radius increases, mass changes from outside the region leak into the estimates (leakage error) [e.g., *Swenson et al.*, 2003]. In this study, three averaging radii were considered: 600 km, 800 km, and 1000 km, each representing the half-wavelength of the Gaussian averaging kernel [*Wahr et al.*, 1998]. The degree 2 zonal term Stokes coefficient $C_{2,0}$ was excluded in the computation, because of the relatively large uncertainties associated with this term [*Tapley et al.*, 2004a]. Water storage changes in the Mississippi River basin were directly extracted from the global mass change fields.

[8] The National Oceanic and Atmospheric Administration's (NOAA) Climate Prediction Center's operational global 2.5° 5-day Merged Analysis of Precipitation (CMAP) is the basis for *P*. This product integrates satellite (infrared and microwave) and gauge observations [*Xie and Arkin*, 1997]. Modeled precipitation fields from NOAA's Global Data Assimilation System (GDAS) [*Derber et al.*, 1991] operational atmospheric analyses were used to disaggregate the CMAP fields to 0.25°, 6-hourly resolutions [*Rodell et al.*, 2004].

[9] Daily discharge measurements for the Mississippi River basin were obtained from the U.S. Army Corps of Engineers (T. Rodgers, personal communication, 2003). These were based on river stage observations from the Vicksburg, Mississippi gauging station.

3. Methods

[10] The change in terrestrial water storage was estimated from GRACE as the difference between one roughly-30-day observation and the previous observation. Accordingly, the monthly basin-scale water balance, neglecting groundwater inflows and outflows, was approximated as

$$\underset{2,1}{S} - \underset{1,1}{S} = \sum_{1,1}^{2,1} P - \sum_{1,1}^{2,1} ET - \sum_{1,1}^{2,1} Q, \qquad (2)$$

where *S*, *P*, *ET*, and *Q* are daily values, and the first index represents the GRACE observation period and the second the day number of that period. To be exact, the terms on the right side of (2) would be integrals of instantaneous values. At continental drainage basin scales, it can safely be assumed that surface drainage divides coincide with groundwater flow divides, so that inputs and outputs of groundwater can be ignored. Rewriting (2) for all pairs of days in the two observation periods and summing all equations yields

$$\left[\underset{2,1}{S} + \cdots + \underset{2,N}{S}\right] - \left[\underset{1,1}{S} + \cdots + \underset{1,N}{S}\right] = \left[\sum_{1,1}^{2,1} P + \cdots + \sum_{1,N}^{2,N} P\right] - \left[\sum_{1,1}^{2,1} ET + \cdots + \sum_{1,N}^{2,N} ET\right] - \left[\sum_{1,1}^{2,1} Q + \cdots + \sum_{1,N}^{2,N} Q\right], \qquad (3)$$

where *N* is the number of days per observation period. After dividing both sides by *N* and simplifying, (4) becomes

$$\Delta\overline{S} = \frac{1}{N} \sum_{n=1}^{N} \sum_{d=D_1+n}^{D_2+n-1} (P_d - ET_d - Q_d), \qquad (4)$$

where $\Delta\overline{S}$ is the change in average water storage, *n* is the day number of the observation period, *d* is the date, and *D* is the first date of the observation period denoted by the index 1 or 2. Equivalently,

$$\Delta\overline{S} = \overline{P} - \overline{ET} - \overline{Q}, \qquad (5)$$

where $\overline{P}$, $\overline{ET}$, and $\overline{Q}$ are running-mean flux accumulations. For example, given two consecutive 30 day periods (60 days), $\overline{ET}$ is the average 30-day evapotranspiration total over all 31 sets of 30 consecutive days within the 60 days. GRACE observation periods are often non-consecutive and different lengths, so that (5) must be expanded to

$$\Delta\overline{S} = \sum_{d=D_1}^{D_1+N_1-1} \frac{d - D_1}{N_1} (P_d - ET_d - Q_d) + \sum_{d=D_1+N_1}^{D_2-1} (P_d - ET_d - Q_d) + \sum_{d=D_2}^{D_2+N_2-1} \frac{D_2 + N_2 - d}{N_2} (P_d - ET_d - Q_d), \qquad (6)$$

where the indices for *D* and *N* denote the observation period. In order to present a daily flux rate, (6) was solved for $\overline{ET}$ (i.e., the sum of all the *ET* terms) and divided by the effective number of days, $\overline{N}$, contributing to the running mean accumulation, where

$$\overline{N} = \{[N_1 - 1]/2\} + \{[D_2 - (D_1 + N_1)]\} + \{[N_2 + 1]/2\}. \qquad (7)$$

5. Summary

[17] Equations were developed for estimating evapotranspiration (*ET*) using a water balance approach with terrestrial water storage changes derived from GRACE and observation based precipitation and runoff. GRACE water storage change estimates were shown to compare favorably with results from a land surface model and a combined atmospheric-terrestrial water balance approach. *ET* was estimated for 14 time periods over the Mississippi River basin, and generally lay in the middle of estimates from a land surface model and two operational atmospheric analysis systems. Uncertainty in GRACE based *ET* was estimated to be 0.86 mm/day. Biases in the model estimates were consistent over time and on the same order as the GRACE uncertainty, so that it was concluded that the technique is valuable for assessing modeled *ET*.